

LECTURE NOTE

SOS 314

INTRODUCTION TO PEDOLOGY AND SOIL PHYSICS

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PART A: PEDOLOGY

WHAT IS A SOIL?

Soil, in terms of its morphological characteristics, is defined as unconsolidated surface material forming natural bodies made up of mineral and organic materials and the living matter within them. Soil is a dynamic entity with material continually and simultaneously added, removed, and transformed. Its formation begins with a parent material derived from either the underlying rock or material transported from somewhere else to its present site.

It is mainly the combined effects of climate and living matter that convert a material to a soil. For example, in temperate rainy environments, moisture and dense vegetation may lead to deep, richly organic soils. In deserts, with the lack of moisture and subsequent vegetation, soils may be thin and remain highly mineral. Human disturbances, such as dwellings, agricultural practices, grave sites, and garbage dumps, may also affect soils, giving them other unique characteristics.

CONSTITUENTS OF SOIL

Soil material has four basic constituents: mineral or inorganic matter, organic substances, water, and air.

MINERAL MATTER

Mineral or inorganic matter can be crystalline or amorphous. A crystal is a chemical compound with a definite chemical formula and a distinct molecular structure. For example, the mineral gibbsite has the chemical formula $\text{Al}(\text{OH})_3$. The small Al^{3+} ion is in the center, surrounded by three hydroxyls (OH) that are equidistant apart. Hydrogen ions can be removed, opening bonds. This allows the crystal to grow laterally and vertically forming sheets that stack up like a deck of cards.

Amorphous minerals lack a repeating long-range structure, but often atoms appear in a definite ratio. A charge may be associated with the surface of mineral matter. Predominantly negative, the charge can also be positive. The type and amount of charge give minerals certain characteristic properties, such as shrink/swell potential and nutrient retention.

ORGANIC SUBSTANCES

Organic substances are molecules with carbon-to-carbon bonds. In soils, these molecules are formed by biochemical activity. Animals, insects, and soil micro-organisms act together to decompose dead plant leaves, root tissue, and animal remains in the soil. Organic matter in soils ranges from leaf litter, where decomposition is minimal and plant species are still recognizable, to a highly decomposed substance called humus, which gives soils a dark brown color.

Organic matter tends to accumulate near the surface where high biological activity, such as leaf litter, roots, and insect life, occurs. As a result, soils near the surface are normally darker in color than the soil horizons a few centimeters below. Organic matter provides a reserve source of plant nutrients and buffers soil against pH changes. It forms very weak cement that, when acting alone, binds soil particles together in a crumb-like structure.

Living organisms residing within the matrix are considered part of the soil. Different fungi, bacteria, protozoa, algae, and actinomycetes play a vital role in converting parent material into soil. Plant roots, rodents, worms, insects, and other burrowing creatures help redistribute matter within a soil profile.

WATER

Soil pores provide an important reservoir for water and atmospheric gases. Soil water is the medium through which nutrients are transferred to plants. Since water has a great capacity to adsorb heat, it can insulate soil from rapid temperature changes. A moist soil is slower to heat in the spring and slower to freeze as the temperature drops.

Hydrogen ions from both organic and inorganic matter dissociate in water, resulting in the soil's pH. A soil's pH affects the solubility of minerals. Soil water may be lost in several ways:

- as transpiration from plant leaves;
- through evaporation from the soil surface;
- by draining through soil pores to groundwater reservoirs;
- through lateral flow; and
- by being held in relatively small pores.

The maximum amount of water a soil holds against gravity is its field capacity. A soil's field capacity is a function of the volume of pores small enough to hold water against gravity. The process is similar to that of a sponge holding water.

AIR

Pore space not filled with water contains gases in concentrations comparable to those in the atmosphere. The soil air is the source of oxygen for root and microbial respiration. A high respiration rate, coupled with the twisting path that a gas must follow in order to diffuse out of soil pores, results in a carbon dioxide concentration about one hundred times greater than that in the atmosphere. Individual gasses move into and out of soil pores primarily by diffusion. After a heavy rain, soil pores fill with water displacing the air.

Oxygen diffuses very slowly through water. Therefore, once a soil becomes saturated with water, respiration quickly removes oxygen from the pores. If the soil layer remains wet for significant periods during the year, the low oxygen content will result in a change in the oxidation/reduction state. Soils so affected become increasingly reduced. Iron oxide minerals in this environment of reduction will change color from red/yellow to a light gray. This change in color can be indicative of a seasonal high water table.

FACTORS OF SOIL FORMATION

The five soil forming factors are: parent material, topography, climate, biological activity, and time. Soil formation begins with a parent material derived from weathering of either the native rock or material transported to the site. The concerted effect of climate and biological activity then transforms parent material by producing the physical and chemical energy to alter minerals and vertically redistribute material through the soil profile. The effect of climate and biological activity is modified by topography.

For example, slope affects the amount of water flowing down through the profile as opposed to running off the surface. Finally, soil forming processes work slowly over time. The intensity and direction of these processes can also change over time. During any given period, one process may dominate; but, with time, another process can become dominant.

PARENT MATERIAL

Parent material is the initial mineral substance that forms a soil. It is derived from the weathering of rocks and may reside at the site of its origin or be transported from somewhere else to its current location. A soil formed from parent material found at the site of its origin is called a residual or sedentary soil. Bedrock weathering in place produces a stony, massive material called saprolite. As physical and some chemical weathering occur, the saprolite becomes denser than

the underlying bedrock. The texture and original rock structure remain, but the material is soft enough to dig with a hand shovel. As chemical weathering converts primary minerals to secondary minerals, particles are redistributed vertically. As material is both added and removed, a soil develops. A residual soil will retain many of the characteristics of the underlying bedrock. Also the texture, mineralogy, pH, and other characteristics of the soil may be a direct result of the saprolite below.

Material can be eroded from one place and transported to another where it becomes parent material for a soil at the new site. Often weathering occurs before the material is transported to the new site. In this case, the soil may have few features in common with the underlying rock. Transported material can bury an existing soil at the new site. Once a depositional episode is completed, time zero for the new soil's formation begins. Several forces can supply energy for the transportation of parent material: ice, wind, water, and gravity.

Ice

Glacial deposits occur at the front and sides of advancing ice. Normally this material is poorly sorted with respect to particle size. Because ice melts from the bottom, this is also true of material deposited under a glacier. Also, material can be deposited as outwash in the glacier's melt-water.

Soils formed from glacial deposits vary in composition depending on the rock type over which the glacier traveled. Since glaciers advance and retreat with time, the composition and depositional environment of the parent material can be quite complex.

Overall, the texture of soil produced in glacial deposits reflects the mode and distance of transport and the type of rock scoured. Shale and limestone scouring tends to produce a soil with relatively more clay and silt-sized material. Igneous and metamorphic rocks produce mostly sandy soils. Deposits beneath the ice usually result in finer textured, denser materials, whereas outwash and front and side deposits are generally coarser.

Wind

Wind deposits two major types of material: eolian sands and loess. Clay-sized material (< 0.002 mm) tends to bind together in aggregates too large to be eroded by wind.

Eolian sands are windblown deposits of material predominantly greater than 0.05 mm (0.05 to 2 mm) in diameter. Most of this material moves in a series of short-distance jumps called saltation. Eolian deposits may move several kilometers from the source.

Material adhering to saltating sand particles and material deposited as an aerosol are the sources of clay in eolian sand. Normally this material has a narrow textural range and is deposited on the leeward side of valleys or bodies of water.

Loess, which is windblown silt-sized material (0.002 to 0.05 mm), once airborne, can travel several hundred kilometers before deposition. The texture of loess usually does not vary in a vertical direction, but tends to thin with horizontal distance from the source.

Windblown material tends to have sharp edges, a conchoidal shape, and surface etching. In contrast, material deposited by water tends to have rounded edges and a polished surface. Careful observation under a hand lens can shed light on the environment present at deposition.

Water

An alluvial or stream-borne deposit occurs in floodplains, fans, and deltas. Because fast-moving water picks up debris, a river meandering downstream will undercut the outer bank of each bend. Water moves slower around the inner bank than the outer bank and therefore loses energy. Thus, coarse material settles out, forming a bar over the inner bank. As water levels rise during floods, the stream overflows its channel and spills over onto the floodplain.

Typically, alluvial deposits are characteristic of the decrease in energy during deposition. Where the stream overflows its bank, the energy is still relatively high; only deposits of coarse material occur, forming a levee. On the far side of a levee, moderate energy is available, and silty material settles.

On the floodplain, water velocity and its corresponding energy is low, and clay settles. Because bars form under moderate energy, this type of sorting does not occur on the plain. However, a floodplain may surround a bar. As the distance from the channel increases, the material's texture becomes finer, and the thickness of the deposit decreases.

Alluvial fans form where water in a channel, carrying sediments downhill, experiences an abrupt reduction in slope. The stream energy is reduced quickly, and material settles. This also occurs where a narrow valley opens onto a wide flat. Fans have a cone shape, widening in the downslope direction. Channels shift easily in fan deposits, and sediments are reworked over

time. The texture of a fan becomes finer with distance from its apex. Normally fans in humid areas are not as steep and cover a much larger area than those in arid regions.

Marine and lacustrine deposits form in low-energy environments under inland seas and lakes. These sediments are typically coarse near the shore and finer toward the middle of the lake or sea. Several shoreline features can be associated with inland water bodies, including deltas, sand dunes, and beaches. Deltas are essentially alluvial fans with their sediments deposited underwater. As lakes dry, evaporite minerals form. Under other conditions, eolian sediments can fill in the lakebed. Such soils have a finer texture and occupy lower sections on the landscape.

Soils formed in shoreline deposits have a coarser texture and occupy higher landscape positions. In lakebeds with a very low influx of sediments, organic substances dominate the sediments, and peats form.

Gravity

Colluvium or hillslope sediments result from the force of gravity and runoff moving downslope. This material may be deposited in catastrophic events, such as mudslides, or by very slow but persistent processes, such as slope wash or surface creep. As viewed from the crest of a hilltop, sediments thicken, and the clay content increases on the downslope.

TOPOGRAPHY (RELIEF)

Topographic relief, or the slope and aspect of the land, has a strong influence on the distribution of soils on a landscape. Position on a slope influences the soil depth through differences in accumulation of erosional debris. Slope affects the amount of precipitation that infiltrates into soil versus that which runs off the surface. Aspect, or the direction a slope is facing, affects soil temperature. In northern hemisphere sites, south-facing slopes are warmer than those facing north. Differences in moisture and temperature regimes create microclimates that result in vegetational differences with aspect. Differences in weathering, erosion, leaching, and secondary mineral formation also can be associated with relief.

CLIMATE

Climate arguably has the greatest effect on soil formation. It not only directly affects material translocation (leaching or erosion, for example) and transformation (weathering), but also indirectly influences the type and amount of vegetation supported by a soil. Precipitation is the main force in moving clay and organic matter from the surface to a depth within the profile.

When a soil is at field capacity, the addition of more water will result in drainage either downward or laterally. Drainage water carries with it dissolved and suspended clay particles that collect at a new location within the soil profile. As a result, soils often show an increase in clay with depth as wind erosion selectively removes clay (and organic matter) from surface horizons. Temperature and moisture affect physical and chemical weathering. Diurnal and seasonal changes in temperature cause particles to expand and contract unevenly, breaking them apart. Heat and moisture are active agents of chemical weathering, the conversion of one mineral into another.

Climate affects the type and amount of vegetation in a region. A warm, humid climate produces the most vegetative growth; however, microbial decomposition is also rapid. The net effect is that tropical and subtropical soils are generally low in organic content. In contrast, organic matter tends to be highest in a cool damp environment where decomposition is slow.

Temperature and the amount of water moving through a profile affects all of the following:

- the amount and characteristics of organic matter;
- the depth at which clay accumulates;
- the type of minerals present;
- soil pH (humid climates tend to produce more acidic soil than do arid climates);
- soil color;
- iron, aluminum, and phosphorus distributions within a soil profile; and
- the depth to calcium carbonate and/or salt accumulation.

BIOLOGICAL ACTIVITY

Biological activity and climate are active forces in soil formation. Soil pedogenesis involves a variety of animals, plants, and microorganisms. Ants, earthworms, and burrowing animals, for example, mix more soil than do humans through plowing and construction. Plant roots remove mineral nutrients from subsoil and redeposit them at the surface in leaf litter. Growing roots open channels through soil where rainwater can wash clay and organic matter down along these channels. Soil microbes decompose plant and animal debris, releasing organic acids. This biochemical activity is the catalyst for a great deal of the oxidation/reduction and other chemical reactions in soil.

The distribution of organic matter in a forest soil is different from that of grassland. The surface soils of forests tend to have concentrated organic matter, which quickly decreases with depth. Grassland soils tend to accumulate organic matter to a greater depth than do forest soils. It is important for pedologists to note that the dark staining from the humic fraction of organic matter can persist in a buried soil. Thus, ancient buried surface soils may be recognized in the field by color alone.

The distribution of iron and aluminum throughout a profile also differs between forest and grassland soils. In forests, due to the greater rainfall, clays and organics drain downward, leaving behind resistant minerals. As a result, iron and aluminum in B horizons in forest soils are found in higher concentrations than in grassland soils.

TIME

Soils develop over time. Soil formation is a dynamic process, where a steady state is slowly approached but only rarely reached. The rate at which a soil forms is related more to the intensity of other soil-forming factors than to chronological age.

Soil development begins with a parent material that has a surface layer altered by vegetation and weathering. For example, a young Coastal Plain soil has relatively uniform material throughout, and is altered only by a dark-stained surface layer that has been formed by vegetation. A more mature soil, on the other hand, shows evidence of the removal and transport of surface-layer clay to a subsurface layer called the B horizon. In an even older soil, chemical weathering and leaching have removed silicon, causing a change in the suite of clay minerals. A senile soil is excessively weathered and dominated by very resistant iron and aluminum oxide minerals. The rate that a young Coastal Plain soil becomes a senile soil depends not on its chronological age but on how rapidly minerals are transported and transformed within the profile.

Human activity frequently alters the process of pedogenesis. Once human activity ends, soil formation can continue as before if no radical change in the soil-forming factors occurred in the interim. Because fine material leaches selectively faster than coarse material, differences between human-altered and undisturbed soils in the ratio of fine to coarse clay may be apparent in a relatively short span of time (one hundred years in a humid environment).

WEATHERING

Weathering is the physical and chemical processes by which rocks and minerals are disintegrated, decomposed, and re-synthesized into new compounds. (Here rocks refer to unconsolidated material and soil at the surface [regolith], while minerals are inorganic substances with a definite chemical structure and formula.)

Weathering encompasses both physical and biogeochemical processes, which generally occur simultaneously. At different times, however, one process may dominate. In a soil forming from saprolite, for example, physical weathering dominates initially. As more surface area is exposed with smaller particles, and as biological activity increases, chemical weathering takes over.

PHYSICAL WEATHERING

Physical weathering is the mechanical disintegration of rocks and minerals into smaller sizes. Some of the several mechanisms that work to break apart rocks include: temperature, water, ice, glaciers, erosion, wind, plants and animals.

Temperature

Seasonal and even day-to-night temperature changes can cause rocks to heat and cool unevenly. As rocks heat up, they expand; as they cool down, they contract. The outer surface expands and contracts faster than the interior, causing the outer surface to separate and peel off.

Water

The force of raindrops beating down on soft rocks, and the scouring effect of suspended material in water flowing over rocks can wear the rocks away with time.

Ice

Water can infiltrate the cracks and pores of rocks and freeze. As the ice expands and thaws, the rocks break up.

Glaciers

Glaciers weather rocks in several ways. The weight of a glacier can crush rocks. As it moves over an area, a glacier can grind and pulverize rocks. As it recedes, the pressure release can cause rocks to expand and crack.

Erosion

Erosion causes pressure-release related weathering.

Wind

Wind suspends fine particles. As the particles are pushed and bounced over one another, they abrade the rock surfaces over which they pass, slowly wearing the rocks down. Over time, the material removed results in pressure-release weathering similar to that of retreating glaciers.

Plants and Animals

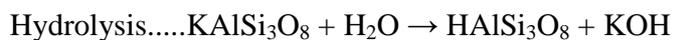
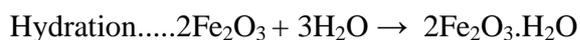
The expansion and decomposition of roots growing in soil can alter the density and coherence of particles. The digging and burrowing of animals can have the same effect.

CHEMICAL WEATHERING

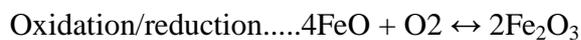
The process of chemical weathering changes the atomic makeup of a mineral. Near the surface, water and biological activity play important roles in chemical weathering. Given time and enough water moving through a profile, even seemingly insoluble minerals will slowly dissolve. These minerals lose a portion of their atomic makeup and re-precipitate as new minerals in a leachate.

Water hydrates minerals, weakening them as it expands the size of their crystals. Hydrolysis removes atoms (ions) from certain minerals and, in the process, splits water molecules affecting the soil pH. Carbon dioxide mixed with water causes a form of acid hydrolysis called carbonation.

Another mechanism of chemical weathering is oxidation/reduction, or the transfer of electrons from one substance to another. Oxidation/reduction affects both the solubility and stability of minerals. Some mechanisms of chemical weathering include:



oxidation / reduction



BUILDING A SOIL

Soil formation is a dynamic process with material continually added, transformed, and/or removed. Beginning with soil from a relatively uniform parent material, windblown sediments and annual floods, for example, add new material to the surface. Physically weathered saprolite adds material to the bottom of the profile. Dissolved and suspended material can be deposited or

redistributed within a soil profile by water flowing below the surface. Evaporite minerals commonly accumulate in the subsoil at the top of a water table. Developmentally, a high annual sediment input often characterizes young soils.

At the other extreme, material is continually removed from a soil. Erosion by wind and water, even under dense vegetation, can remove five tons of soil per acre, per year. Whenever precipitation exceeds the field capacity of a soil, material can be leached below the soil solum (the root zone or an area active in soil pedogenesis). Biochemical degradation can remove organic matter. This can lead to a significant reduction in soil volume.

Changes also occur within a soil profile. Material is converted from one form to another and translocated within a profile. Clay, organic matter, and iron/aluminum ions typically migrate out from surface horizons especially in humid-region soils. The root zone provides a good environment for biogeochemical activity. Two examples of this are melanization and gleization. Melanization is the darkening of a soil layer by organic matter. This process gives the surface (A horizon) its brownish color. Gleization is the reduction of mostly iron-bearing minerals. It produces a gray to greenish color in soil. Saturation of a soil layer for long periods within a year usually causes this condition.

MINERALS IN SOIL ENVIRONMENT

A. Composition of the Earth's Crust

Most of the weight and volume of the earth's crust is made up by only a few elements (Table 1). Oxygen and Si make up most of the weight, while oxygen alone accounts for more than 90% of the volume. In general, the larger O atoms are in an approximately close-packed arrangement held together by smaller metal atoms in the interstitial space. Most of the elements in the crust and in soils occur in minerals. Thus, the elements listed in Table 1 are major constituents of the minerals found in the soil environment.

Table 1. The twelve most common chemical elements in the earth's crust (Klein and Hurlbut, 1993).

Element	Crustal average	Mole fraction gkg ⁻¹	Ionic radius (nm)	Volume %
O	466	0.6057	0.136 (3)	92.88
Si	277.2	0.2052	0.026 (4)	0.22
Al	81.3	0.0626	0.039 (4)	0.23
Fe	50	0.0186	0.078 (6)	0.54
Ca	36.3	0.0188	0.1 (6)	1.15
Na	28.3	0.0256	0.102 (6)	1.66
K	25.9	0.0138	0.151 (6)	2.89
Mg	20.9	0.0179	0.072 (6)	0.41
Ti	4.4	0.0019	0.061 (6)	0.03
H	1.4	0.0289	?	?
P	1	0.0007	0.017 (4)	0
Mn	0.9	0.0003	0.083 (6)	0.01

Numbers in parentheses refer to coordination number. Radii for Fe and Mn are for the reduced (²⁺) form. ? Ionic radius and volume of H⁺ are negligible compared to O₂

Soil minerals are also referred to as either *primary* or *secondary* minerals. Primary minerals form at elevated temperatures. They are usually derived from igneous or metamorphic rocks, but they can be inherited from sedimentary rocks as well. Secondary minerals are formed by low temperature reactions and are either inherited from sedimentary rocks or formed in soils by weathering (Jackson, 1964). The separation of minerals into primary and secondary mineral classes is not mutually exclusive and some minerals can occur in both. The concept is useful however, and appears widely in the soil science literature.

Although different classification schemes could be used, mineralogists have determined that first separating minerals into groups based on their chemical composition gives classes with the greatest similarities in many other properties. Thus, minerals are first divided into classes depending upon the dominant anion or anionic group. The classes include: (1) native elements, (2) sulphides, (3) sulphosalts, (4) oxides and hydroxides, (5) halides, (6) carbonates, (7) nitrates, (8) borates, (9) phosphates, (10) sulphates, (11) tungstates, and (12) silicates (Klein and Hurlbut, 1993). These classes are then subdivided based on chemical and structural similarities.

Mineral classes such as the native elements, sulphosalts, nitrates, borates, and tungstates occur only rarely in soils.

The major mineral classes represented in soils are discussed below.

1. Halide, Sulphate, and Carbonate Minerals

The major soil minerals of this group are halite (NaCl), gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$), calcite (CaCO_3), and dolomite [$\text{CaMg}(\text{CO}_3)_2$] (Table 2). This group is characterized by minerals with relatively simple structures. The halite structure is one of the simplest of all minerals. It consists of

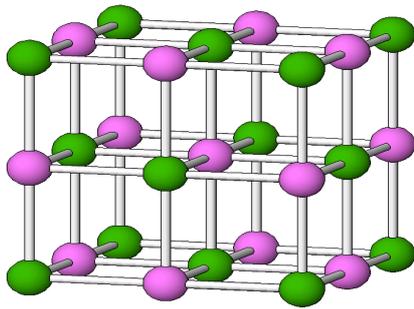


Fig. 1. Structure of Halides

alternating Na^+ and Cl^- ions arranged in cubic closest packing (Fig. 1). The other minerals in this group have similar structures with cations such as Ca^{2+} , Mg^{2+} , or Fe^{2+} alternating with anions such as S_2^{2-} , SO_4^{2-} , or CO_3^{2-} . The bonds between the cations and anions are predominantly ionic. These minerals are among the most soluble and the softest of all soil minerals and they are easily broken down by physical and chemical weathering processes. They occur mainly in soils of arid regions or in youthful soils in more humid regions, where weathering has been minimal.

Halite is the most soluble of this group and accumulates in only the most arid environments. It is one of the minerals present in the salic horizon of Aridisols. Gypsum is about 100 times less soluble than halite, but it too is abundant only in arid regions. Gypsum is a major mineral in the gypsic horizon of Aridisols. Halite and gypsum could also occur in soils that have been contaminated by salt water, or by soluble salts leaching from industrial stock piles or waste piles. Calcite and dolomite are common carbonate minerals that occur in a variety of soils. These minerals precipitate in the soil profile in arid and semiarid climates (aridic and some ustic soil moisture regimes). Calcic and petrocalcic horizons form if the accumulation is great enough. Carbonate minerals in many soils are inherited from limestone or other calcareous parent materials. Carbonates are usually stable and can be found throughout the soil profile under aridic

and ustic soil moisture regimes (arid to subhumid climates), but are leached from the soil profile and are generally present only in the C horizons under the udic soil moisture regime (humid climates). Calcite and dolomite can be introduced into soils originally free of these minerals via the limestone aggregate used for road construction in some areas.

2. Sulphides

Pyrite, FeS_2 , the most common mineral in this group, does not occur extensively in soils, but when it is present it causes some unique problems. Pyrite precipitates in soils on wet tidal flats and river deltas of some coastal areas and also occurs in some geologic formations originally deposited in similar environments. Thus, pyrite often occurs in close association with coal. Pyrite is unstable under oxidizing conditions and weathers quickly when pyritic soils are drained or when mining leaves pyritic material on the surface. The weathering products include the minerals jarosite, $\text{KFe}_3(\text{SO}_4)_2(\text{OH})_6$, and gypsum and sulfuric acid, H_2SO_4 . The large amount of acidity causes problems in utilizing pyrite-containing soils and in re-vegetating mined areas and is a serious and costly environmental problem. Such soils are usually referred to as acid-sulphate soils.

3. Oxides, Hydroxides and Oxy-hydroxides

Primary minerals break down during weathering and release cations and anions that recombine to form other more stable minerals. Several elements, in particular Al, Fe, and Mn, form oxide, hydroxide, or oxy-hydroxide minerals that are stable in the soil weathering environment. Representative mineral species are given in Table 2.

a. Aluminum

Gibbsite, $[\text{Al}(\text{OH})_3]$ is the most common Al hydroxide mineral in soils. It is generally associated with the latter stages of weathering when leaching of silica has progressed to the point that phyllosilicate minerals no longer form. Gibbsite is common in highly weathered Oxisols of tropical regions. It has a very low cation exchange capacity and contributes to the low native fertility of most Oxisols. Gibbsite is also commonly found at the weathering interface between igneous rock and saprolite and in Andisols formed from volcanic ash. Despite the high solution silica content in these environments, gibbsite precipitates as a metastable phase that eventually dissolves and gives way to more silica-rich minerals.

b. Iron

Iron oxide minerals form from Fe released from primary minerals. Iron oxides are strong pigments and small amounts of these minerals account for most of the brown and red colors of soils. Goethite (FeOOH) is the most common mineral of this group. It accounts for the brownish to yellowish color of many soils, although it may be present in only small quantities.

Hematite (Fe₂O₃) is only slightly less common than goethite and usually occurs in association with it. Hematite is usually bright red and is responsible for the red color of many soils. Goethite and hematite are stable minerals in an oxidizing environment. Large amounts of these two minerals in well-aerated soils, usually in association with gibbsite and kaolinite, usually indicate an advanced stage of weathering. In soils that are saturated with water for at least some time during the year, the very insoluble Fe³⁺ in goethite, hematite, and other iron oxides can be reduced to very soluble Fe²⁺.

The Fe²⁺ can easily move with the soil water to other parts of the soil profile, or even to other associated soils in the landscape, where it can then re-oxidize to Fe³⁺ and re-precipitate as goethite, lepidocrocite, or ferrihydrite when oxidizing conditions return. Repeated cycles of oxidation and reduction give rise to mottles and nodules that reflect an inhomogeneous distribution of iron oxide minerals within a soil. Soil scientists use these inhomogeneous or

Table 2. Common non-silicate minerals in soils.

Mineral class	Mineral	Chemical formula
Halides	Halite	NaCl
Sulfates	Gypsum	CaSO ₄ •2H ₂ O
	Jarosite	KFe ₃ (SO ₄) ₂ (OH) ₆
Sulfides	Pyrite	FeS ₂
Carbonates	Calcite	CaCO ₃
	Dolomite	CaMg(CO ₃) ₂
	Nahcolite	NaHCO ₃
	Trona	Na ₂ CO ₃ •NaHCO ₃ •2H ₂ O
	Soda	Na ₂ CO ₃ •10H ₂ O
Oxides and Hydroxides		
Aluminum	Gibbsite	Al(OH) ₃
Iron	Hematite	Fe ₂ O ₃
	Goethite	FeOOH
	Lepidocrocite	FeOOH
	Maghemite	Fe ₂ O ₃

	Ferrihydrite	$\text{Fe}_5\text{O}_7(\text{OH})\cdot 4\text{H}_2\text{O}$
	Magnetite	Fe_3O_4
Manganese	Birnessite	$(\text{Na}, \text{Ca}, \text{Mn}^{2+})\text{Mn}_7\text{O}_4\cdot 2.8 \text{H}_2\text{O}$
	Lithiophorite	$\text{LiAl}_2\text{Mn}_2^{4+}\text{Mn}^{3+}\text{O}_6(\text{OH})_6$
	"Hollandite	$\text{Ba}(\text{Mn}^{4+}, \text{Mn}^{3+})_8\text{O}_{16}$ "
	"Todorokite	$(\text{Na}, \text{Ca}, \text{K})_{0.3-0.5}(\text{Mn}^{4+}, \text{Mn}^{3+})_6 \text{O}_{12.3}\cdot 5\text{H}_2\text{O}$
Titanium	Rutile	TiO_2
	Anatase	TiO_2
	Ilmenite	$\text{Fe}^{2+}\text{TiO}_3$

(Klein and Hurlbut (1993), Kampf et al. (1999))

mottled color patterns to estimate the depth to a soil water table and the length of time the soil is saturated during the year. Thus, when delineating wetlands, determining the best location for a septic system leach field, or mapping soils, soil scientists rely on predictable soil color patterns that result from iron oxide mineralogy and distribution.

c. Manganese

Manganese oxides and hydroxides (Table 2) are commonly found in soils as brown or black nodules or as thin coatings on the faces of soil structural units. They are often associated with Fe oxides. Manganese occurs frequently as birnessite or lithiophorite in soils. Manganese can be oxidized and reduced in the soil environment similar to iron. Thus, the inhomogeneous distribution of Mn into nodules is an indicator of reduction and oxidation as a result of periodic water saturation.

d. Titanium

Rutile and ilmenite (Table 2) occur in soils mainly as primary minerals inherited from igneous rocks. Anatase is less common and is generally considered a secondary mineral. Although frequently found in soils, these minerals do not occur in sufficient quantity to impact soil physical or chemical properties.

4. Silicates

The silicate mineral class is an extremely large and important group of minerals. Nearly 40% of the common minerals are silicates, as are most minerals in igneous rocks. Silicates constitute well over 90% of the earth's crust (Klein and Hurlbut, 1993) and comprise the bulk of most soils as well.

Silicates occur as both primary minerals inherited from igneous or metamorphic rocks and as secondary minerals formed from the weathering products of primary minerals. As explained in more detail below, the fundamental unit of all silicate structures is the SiO₄ tetrahedron. It consists of four O²⁻ ions at the apices of a regular tetrahedron coordinated to one Si⁴⁺ at the center. The individual tetrahedra are linked together by sharing O²⁻ ions to form more complex structures. Several different arrangements of the SiO₄ tetrahedra occur, partly accounting for the

Table 3 Classification of silicate minerals.

Silicate class, Unit composition, Arrangement of SiO ₄ tetrahedra	Mineral	Ideal Formula?
Nesosilicates(SiO ₄) ⁴⁻	Olivine	(Mg,Fe) ₂ SiO ₄
	Forsterite	Mg ₂ SiO ₄
	Fayalite	Fe ²⁺ SiO ₄
	Zircon	ZrSiO ₄
	Sphene	CaTiO(SiO ₄)
	Topaz	Al ₂ SiO ₄ (F,OH) ₂
	Garnets	X ₃ Y ₂ (SiO ₄) ₃ , X=Ca,Mg,Fe ²⁺ ,Mn ²⁺ , Y=Al,Fe ³⁺ ,Cr ³⁺
	Andalusite	Al ₂ SiO ₅
	Sillimanite	Al ₂ SiO ₅
	Kyanite	Al ₂ SiO ₅
Staurolite	Fe ₂ Al ₉ O ₆ (SiO ₄) ₄ (O,OH) ₂	
Sorosilicates(Si ₂ O ₇) ⁶⁻ (Si ₂ O ₇)(OH)	Epidote	Ca ₂ (Al,Fe)Al ₂ O(SiO ₄)
Cyclosilicates(Si ₆ O ₁₈) ¹²⁻	Beryl	Be ₃ Al ₂ (Si ₆ O ₁₈)
	Tourmaline	(Na,Ca)(Li,Mg,Al)(Al,Fe, Mn) ₆ (BO ₃) ₃ (Si ₆ O ₁₈)(OH) ₄
Inosilicates (single chains)(SiO ₃) ₂	Pyroxenes	
	Augite	(Ca,Na)(Mg,Fe,Al)(Si,Al) ₂ O ₆
	Enstatite	MgSiO ₃
	Hypersthene	(Mg,Fe)SiO ₃
	Diopside	CaMgSi ₂ O ₆
	Hedenbergite	CaFeSi ₂ O ₆
	Pyroxenoids	
	Wollastonite	CaSiO ₃
Rhodonite	MnSiO ₃	
Inosilicates (double chains)Si ₄ O ₁₁) ⁶⁻	Amphiboles	
	Hornblende	(Ca,Na) ₂₋₃ (Mg,Fe,Al) ₅ Si ₆ (Si,Al) ₂ O ₂₂ (OH) ₂
	Tremolite	Ca ₂ Mg ₅ Si ₈ O ₂₂ (OH) ₂

	Actinolite	$\text{Ca}_2(\text{Mg,Fe})_5\text{Si}_8\text{O}_{22}(\text{OH})_2$
	Cummingtonite	$(\text{Mg,Fe})_7\text{Si}_8\text{O}_{22}(\text{OH})_2$
	Grunerite	$\text{Fe}_7\text{Si}_8\text{O}_{22}(\text{OH})_2$
Phyllosilicates (Si_2O_5) ²⁻	Micas	
	Muscovite	$\text{KAl}_2(\text{AlSi}_3\text{O}_{10})(\text{OH})_2$
	Biotite	$\text{K}(\text{Mg,Fe})_3(\text{AlSi}_3\text{O}_{10})(\text{OH})_2$
	Phlogopite	$\text{KMg}_3(\text{AlSi}_3\text{O}_{10})(\text{OH})_2$
	Chlorites	$(\text{Mg,Fe})_3(\text{Si,Al})_4\text{O}_{10}(\text{OH})_2$
	Clay minerals (selected)	
	Talc	$\text{Mg}_3\text{Si}_4\text{O}_{10}(\text{OH})_2$
Pyrophyllite	$\text{Al}_2\text{Si}_4\text{O}_{10}(\text{OH})_2$	
Kaolinite	$\text{Al}_2\text{Si}_2\text{O}_5(\text{OH})_4$	
Smectite	$\text{M}^{+}_{0.3}\text{Al}_2(\text{Al}_{0.3}\text{Si}_{3.7})\text{O}_{10}(\text{OH})_2$ $\text{M}^{+} = \text{Ca}^{2+}, \text{Mg}^{2+}, \text{K}^{+}, \text{etc.}$	
Vermiculite	$\text{M}^{+}_{0.7}\text{Al}_2(\text{Al}_{0.7}\text{Si}_{3.3})\text{O}_{10}(\text{OH})_2$ $\text{M}^{+} = \text{Ca}^{2+}, \text{Mg}^{2+}, \text{K}^{+}, \text{etc.}$	
	Serpentines	
	Antigorite	$\text{Mg}_3\text{Si}_2\text{O}_5(\text{OH})_4$
	Chrysotile	$\text{Mg}_3\text{Si}_2\text{O}_5(\text{OH})_4$
Tectosilicates(SiO_2) ⁰	Feldspars	
	Orthoclase	KAlSi_3O_8
	Albite	$\text{NaAlSi}_3\text{O}_8$
	Anorthite	$\text{CaAl}_2\text{Si}_2\text{O}_8$
	SiO ₂ Group	
	Quartz	SiO_2
	Tridymite	SiO_2
	Cristobalite	SiO_2
	Zeolites	
	Analcime	$\text{NaAlSi}_2\text{O}_6 \cdot \text{H}_2\text{O}$
Feldspathoids		
Nephelene	$(\text{Na,K})\text{AlSiO}_4$	

Allen and Fanning (1983).

Klein and Hurlbut (1993).

large number of silicate minerals and providing the basis for their classification. The tetrahedra may be present as single tetrahedra (nesosilicates), double tetrahedra (sorosilicates), rings (cyclosilicates), single or double chains (inosilicates), sheets (phyllosilicates) or three-dimensional frameworks (tectosilicates) (Tabl 3). In most of these arrangements adjacent SiO_4 tetrahedra share corners, that is, they share a common O^{2-} . The more common minerals from each silicate class likely to be found in soils are given in Table 3.

The most common minerals in igneous rocks are the olivines, pyroxenes, amphiboles, micas, feldspars, and quartz. These primary minerals predominate in the sand and silt size fractions of soils. The clay (<2 mm) fraction of most soils is dominated by phyllosilicate minerals.

SOIL PROFILE AND PROFILE DESCRIPTION

MASTER HORIZONS

Horizontal layers of soil called horizons can be described by their different morphological characteristics. Capital letters designate master horizons, which are further subdivided by Arabic numerals. Master horizons are used to describe similar appearing soil layers and should not be confused with diagnostic horizons used to classify soils.

O Horizon

The O horizon is a surface layer dominated by organic material. An O horizon may be found below the surface if it has been buried. Predominantly found in forested regions, the O horizon is composed of leaf litter in various stages of decay.

A Horizon

The A horizon is the uppermost mineral layer. It may lie below the O horizon. An A horizon has a high concentration of humus and is not dominated by the migration of clay, humus, aluminum, or iron into or out of the horizon. The humus content gives it a darker color than the horizon below.

E Horizon

The E horizon is a layer of eluviation where clay organic matter and iron and aluminum oxides have been leached out. Remaining material tends to be light colored and coarse textured. The E horizon is normally found below an O or an A horizon and above a B horizon. However, it may separate sections of a B horizon.

B Horizon

The B horizon is a subsurface layer showing evidence of one or more of the following processes:

1. illuvial accumulation of alumino-silicate clay, iron, aluminum, gypsum, or silica;
2. carbonate removal;
3. residual concentration of sesquioxides;

4. coating of sesquioxides, which makes the horizon conspicuously lower in color value, higher in chroma, or redder in hue without apparent illuviation of iron than that found in the overlying and underlying horizons;
5. alteration that forms silicate clay or liberates oxides, or both, and that forms a granular, blocky, or prismatic structure if volume changes accompany changes in moisture context;
or
6. brittleness.

C Horizon

The C horizon is a layer of minimal alteration. Material may be similar to or unlike that from which the other horizons formed. C horizons lack the properties of O, A, E or B horizons, and can include coprogenous earth (sedimentary peat), diatomaceous earth, aprolite, unconsolidated bedrock, and other uncemented geologic materials or materials soft enough for excavation with moderate difficulty.

R Layer

An R layer refers to hard bedrock. Material is cemented and manual excavation is impossible. Intrusive soils can be found in rare cracks in the bedrock. Examples of R layer material include: granite, basalt, quartzite, indurated limestone, or sandstone.

TRANSITIONAL HORIZONS

Transitional horizons are dominated by properties of one master horizon but have the subordinate properties of another. These are designated by two capital letters, for example, AB, EB, BE, or BC. The first letter represents the dominant horizon characteristics; the second indicates the weaker expressed characteristics.

A second type of transitional horizon has two distinct parts with recognizable properties of the two master horizons indicated by the capital letters. Parts of one surround the other. This type of transitional horizon is designated by a capital letter for the part with the greatest volume, followed by a slash and another capital letter for the secondary part (for example, E/B, B/E, or B/C).

SUBORDINATE DISTINCTIONS

Master horizons are further divided by subordinate characteristics, which usually do not apply to transitional horizons. Subordinate distinctions are identified by lower-case letters, called suffix symbols. In some cases, they describe an accumulation of material.

This means that the so-designated horizons contain more of the material in question than is presumed to have been present in the parent material. For example, Bt refers to a B horizon with more clay than normal. The symbols and their meanings follow.

- *a* . *highly decomposed organic material*. Used with O to indicate the most highly decomposed organic materials, which have rubbed fiber content of less than 17 percent of the volume.
- *b* . *buried genetic horizon*. Used in mineral soils to indicate identifiable buried horizons with major genetic features that were developed before burial. Genetic horizons may or may not have formed in the overlying material, which may be either like or unlike the assumed parent material of the buried soil. This symbol is not used in organic soils or to separate an organic from a mineral layer.
- *c* . *concretions or nodules*. Indicates a significant accumulation of concretions or nodules. Cementation is required, but the cementing agent is not specific, except that it cannot be silica. The symbol is not used if the concretions or nodules consist of dolomite or calcite, or more soluble salts. It is used if the nodules or concretions are enriched with minerals that contain iron aluminum, manganese, or titanium.
- *d* . *physical root restriction* . Indicates root-restricting layers in naturally occurring or man-made unconsolidated sediments or materials, such as dense basal till, plow pans, and other mechanically compacted zones.
- *e* . *organic material of intermediate composition*. Used with O to indicate organic materials of intermediate composition with rubbed fiber content between 17 and 40 percent (by volume).
- *f* . *frozen soil*. Indicates permanent ice content in a horizon or layer. The symbol is not used for seasonally frozen layers or for so-called dry permafrost (material that is colder than 0° but does not contain ice).

- ***g*** . *strong gleying*. Indicates either that iron has been reduced and removed during soil formation, or that saturation with stagnant water has preserved it in a reduced state. Most of the affected layers have a chroma of 2 or less, and many have redox concentrations. The low chroma can represent either the color of reduced iron or the color of uncoated sand and silt particles from which the iron has been removed. The symbol *g* is not used for materials of low chroma that have no history of wetness, such as some shales or E horizons. If *g* is used with B, pedogenic change in addition to gleying is implied. The horizon is designated Cg if no other pedogenic change besides gleying has occurred.
- ***h*** . *illuvial accumulation of organic matter*. Used with B to indicate the accumulation of illuvial, amorphous, dispersible organic-matter-sesquioxide complexes if the sesquioxide component is dominated by aluminum but is present only in small quantities. The organo-sesquioxide material coats sand and silt particles. In some horizons, these coatings have coalesced, filled pores, and cemented the horizon. The symbol *h* is also used in combination with *s*, as in Bhs, if the amount of sesquioxide component is significant but the color value and chroma of the horizon when moist is 3 or less.
- ***i***. *slightly decomposed organic matter*. Used with O to indicate the least decomposed of the organic materials. Its rubbed fiber content is 40 percent or more (by volume).
- ***k***. *accumulation of carbonates*. Indicates an accumulation of alkaline-earth carbonates, commonly calcium carbonate. It is usually found in arid or dry region.
- ***m***. *cementation or induration*. Indicates continuous or nearly continuous cementation. The symbol *m* is used for horizons that are more than 90 percent cemented, although they may be fractured. The cemented layer is physically root-restrictive. The predominant cementing agent (or the two dominant cementing agents) may be indicated by using defined letter suffixes, singly or in pairs. Following are some suffix combinations and what they indicate:
 - *km* . cementation by carbonates;
 - *qm* . cementation by silica;
 - *sm* . cementation by iron;
 - *ym* . cementation by gypsum;
 - *kqm*. cementation by lime and silica; and

- *zm* . cementation by salts more soluble than gypsum.
- *n* . *accumulation of sodium*. Indicates an accumulation of exchangeable sodium.
- *o* . *residual accumulation of sesquioxides*
- *p* . *tillage or other disturbance*. Indicates a disturbance of the surface layer by mechanical means, pasturing, or similar uses. A disturbed organic horizon is designated Op. A disturbed mineral horizon is designated Ap, even though it is clearly a former E, B, or C horizon.
- *q* . *accumulation of silica*. Indicates an accumulation of secondary silica.
- *r* . *weathered or soft bedrock*. Used with C to indicate root-restrictive layers of saprolite, such as weathered igneous rock, or of soft bedrock, such as partly consolidated sandstone, siltstone, and shale. Excavation difficulty is low to high.
- *s* . *illuvial accumulation of sesquioxides and organic matter*. Used with B to indicate an accumulation of illuvial, amorphous, dispersible, organic-matter-sesquioxide complexes if both organic-matter and sesquioxide components are significant, and if color value and chroma of the horizon when moist is 4 or more. The symbol is also used in combination with the symbol *h*, as in Bhs, if both the organic-matter and sesquioxide components are significant, and if the color value and chroma, moist, is 3 or less.
- *ss* . *presence of slickensides*. Indicates the presence of slickensides. Slickensides result directly from the swelling of clay minerals and shear failure, commonly at angles of 20 to 60 degrees above horizontal. They are indicators that other vertic characteristics, such as wedge-shaped peds and surface cracks, may be present.
- *t* . *accumulation of silicate clay* Indicates an accumulation of silicate clay that has either formed and subsequently been translocated within the horizon or has been moved into the horizon by illuviation, or both. At least some part of the horizon should show evidence of clay accumulation either as coatings on surfaces of peds or in pores, or as lamellae or bridges between mineral grains.
- *v* . *plinthite*. Indicates the presence of iron-rich humus-poor reddish material that is firm or very firm when moist and hardens irreversibly when exposed to the atmosphere and to repeated wetting and drying.

- *w* . *development of color or structure*. Used with B to indicate the development of color and structure, or both, with little or no apparent illuvial accumulation of material. It should not be used to indicate a transitional horizon.
- *x* . *fragipan character*. Indicates a genetically developed layer with a combination of firmness, brittleness, and commonly a higher bulk density than adjacent layers. Some part of the layer is physically root-restrictive.
- *y* . *accumulation of gypsum*
- *z* . *accumulation of salts more soluble than gypsum*

DIAGNOSTIC HORIZONS: EPIPEDONS

These are horizons mainly used for soil classification. Although they are supposed to be identified on the field, many cannot be surely identified on the field particularly in the tropics. Therefore, laboratory analysis are use to confirm and establish the field identification.

There are two types of diagnostic horizons. These are surface (epipedon) and subsurface (master horizon).

EPIPEDONS

Master horizons describe a soil profile, while diagnostic horizons are used to classify soils. Whereas master horizons are based on appearance, diagnostic horizons are based on soil formation processes. These two classification schemes are not complementary. Diagnostic horizons can contain all or part of more than one master horizon.

An epipedon is the surface, or uppermost soil horizon. It may be thinner than the soil profile A horizon, or include the E or part or all of the B horizon. Epipedons derived from bedrock lack rock structure and are normally darkened by organic matter.

Anthropic epipedon

While similar to the mollic epipedon, the anthropic epipedon contains greater than 250 ppm citric acid soluble P_2O_5 with or without a 50 percent base saturation and requires that the soil is moist three months or more over 8 to 10 years. It is commonly found in fields cultivated over long periods of time.

Histic epipedon

This organic horizon is water saturated long enough for reduced conditions to occur unless artificially drained. It is 20 to 60 cm thick and has a low bulk density often less than 1 g cm⁻³. The actual organic matter content is dependent on the percent clay. If the soil has not been plowed, it must contain between 12 percent or more organic carbon with no clay and 18 percent or more organic carbon with 60 percent or more clay. When the soil has been plowed, the organic carbon content is from 8 percent with no clay to 16 percent with 60 percent or more clay.

Mollic epipedon

This epipedon is a soft dark grassland soil. Its organic carbon content is 0.6 percent or more resulting in a color value of 3 or less moist, 5 or less dry. Its base saturation is 50 percent or more. It measures a minimum of 18 cm thick if not directly above a petrocalcic horizon, duripan, or a lithic or paralithic contact, and contains less than 250 ppm P₂O₅. Moist three months or more each year, it cannot have both hard consistence and massive structure.

Ochric epipedon

This epipedon does not meet the definitions of any other surface horizon. It does not have the thickness, percent organic carbon, or color to be a mollic or umbric epipedon. The ochric epipedon extends to the first illuvial (B) horizon.

Plaggen epipedon

This man-made horizon is 50 cm or more thick and has resulted from centuries of accumulation of sod, straw, and manure, for example. It commonly contains artifacts such as pottery and bricks.

Umbric epipedon

Mollic-like in thickness, organic carbon content, color, P₂O₅ content, consistence, and structure, this epipedon has less than 50 percent base saturation.

DIAGNOSTIC SUBSURFACE HORIZONS

Diagnostic subsurface horizons can be categorized as weakly developed horizons, as horizons featuring an accumulation of clay, organic matter, or inorganic salts, as cemented horizons, or as strongly acidic horizons.

Agric

This horizon forms under a plow layer. It normally has lamellae (finger-shaped concentrations of material) of illuvial humus, silt, and clay.

Albic

Clay, humus, and other coatings have been leached from this eluvial horizon, leaving light-colored sand and silt particles.

Argillic

This illuvial horizon of mostly high-charged layer silicate clay has clay films on the faces of peds or some indication of clay movement. It is at least one-tenth the thickness of all overlying horizons. If the overlying horizon has less than 15 percent clay, the argillic has 3 percent more clay than the eluvial horizon above. If the overlying horizon has 15 to 40 percent clay, the argillic has 1.2 times that amount. If the overlying horizon has over 40 percent clay, the argillic has 8 percent more clay.

Calcic

Measuring 15 cm or more thick, this horizon is not indurated or cemented, and has evidence of calcium carbonate movement. It has a 15 percent or more CaCO₃ equivalent unless there is below 18 percent clay, then the requirement is a 5 percent or more CaCO₃ equivalent. If the horizon is cemented, it is classified as petrocalcic.

Cambic

This horizon shows some evidence of alterations but is very weakly developed between A and C horizons. The cambic horizon has less illuviation evidence than found in the argillic and spodic horizons.

Duripan

This subsurface horizon is cemented by silica in more than 50 percent of its volume. It dissolves in concentrated basic solution or altering acid and then basic solutions, but does not slake in HCl.

Fragipan

A fragipan is a brittle horizon situated at some depth below an eluvial horizon. It has a low organic matter content, lower bulk density than overlying horizons, and hard or very hard consistence when dry.

Glossic

This transitional horizon has parts of an eluvial horizon and the remnants of a degrading argillic, kandic, or natric horizon.

Gypsic

An illuvial horizon, the gypsic is 15 cm or more thick with 5 percent or more gypsum and at least 1 percent by volume visible gypsum. If the horizon is cemented, it is classified as petrogypsic.

Kandic

The kandic is a horizon with an illuvial accumulation of 1:1 (kaolinite-like) clay that has a CEC of less than 16 cmolc kg⁻¹ clay. A clay increase within 15 cm of the overlying horizon is 4 percent or more if the surface has less than 20 percent clay; 20 percent or more if the surface has 20 to 40 percent clay; or 8 percent or more if the surface has greater than 40 percent clay. The horizon is 30 cm thick unless there is a lithic, paralithic, or petroferric contact, in which case minimum thickness is 15 cm. Organic carbon constantly decreases with increasing depth.

Natric

The natric horizon is similar to the argillic horizon with the additional characteristics of columnar structure. It has an exchangeable sodium percentage of 15 percent or more.

Oxic

The oxic horizon contains highly weathered clays. It is 30 cm or more thick and has a CEC of less than 16 cmolc kg⁻¹ clay. Less than 10 percent of the minerals are weatherable. Within a distance of 15 cm, there is an increase in clay of 4 percent or less if the surface horizon contains less than 20 percent clay; less than 20 percent if the surface contains 20 to 40 percent clay; or 8 percent or less if the surface contains 40 percent or more clay.

Placic

This subsurface horizon is cemented by iron, iron and manganese, or iron and organic matter.

Salic

Measuring 15 cm or more thick, the salic horizon contains at least 2 percent soluble salt. A 1:1 soil to water extract has an electrical conductivity of 30 dS/m⁻¹ (decisiemens per meter) or more.

Sombric

The sombric horizon has an illuvial accumulation of humus that is not associated with aluminum (spodic) or sodium (natric).

Spodic

This illuvial horizon contains high pH dependent charge material. A sandy-textured horizon, it has an accumulation of humus with aluminum and/or iron.

Sulfuric

The sulfuric horizon forms as a result of draining soil with a high sulfide content that is oxidized to sulfates, drastically reducing the pH. It is at least 15 cm thick and has a pH of 3.5 or less.

BASIC PRINCIPLES OF SOIL CLASSIFICATION

Whenever two objects are compared with one another, either critically or casually, they are being distinguished one from the other, and are in effect, being classified. The nature and extent of the real and significant differences that exist between the two objects will become more or less apparent depending upon the observational skill of the observer and upon their fundamental knowledge of the objects. The process of classification involves the formation of classes by grouping the objects on the basis of their common properties. In any system of classification, those groups about which the greatest number of things can be stated for the chosen objective are generally the best and the most useful classification groupings. A classification system is an organization of the knowledge of the moment.

• Why classify?

We classify soils for the following purposes:

1. To organize our knowledge about soils
2. To provide an organizational chart or map of the world of soils as we perceive it - the soil survey
3. To develop principles and guidelines for proper use and management:
 - a. to predict behavior
 - b. to identify best uses
 - c. to estimate productivity
 - d. to identify potential problems
4. To facilitate easier transfer of information and technology
5. To provide a basis for research and experimentation
6. To understand relationships among individuals of the population

Principles of classification

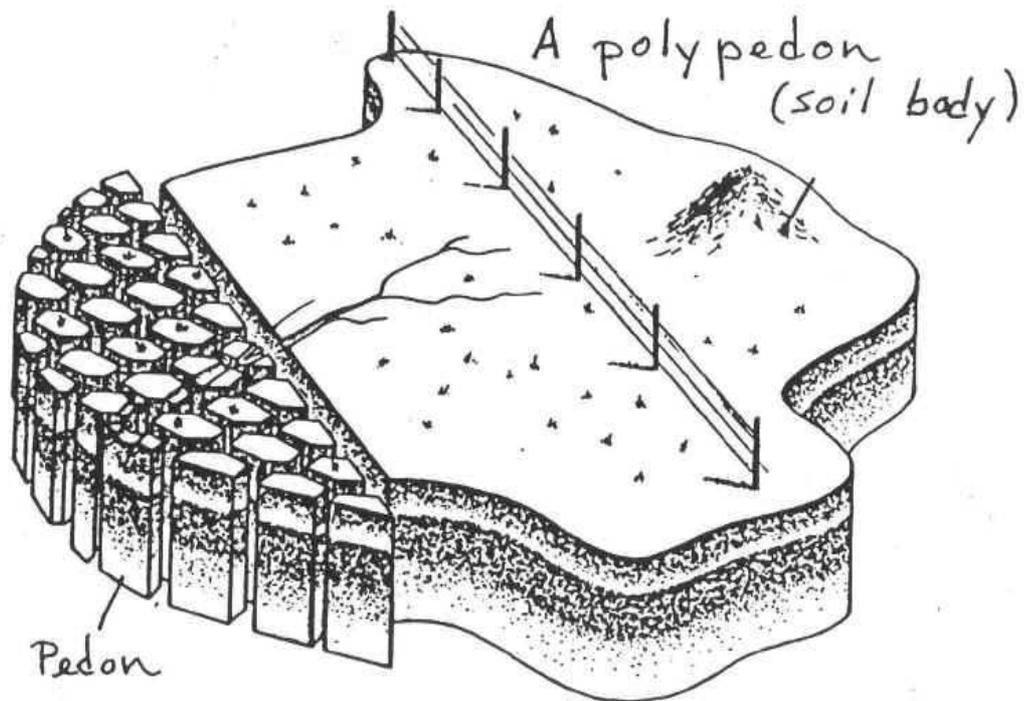
1. Principle of **Purpose**. The reasons for wanting to organize soil knowledge.
2. Principle of **Domain**. The universe of objects relevant to the purpose.
3. Principle of **Identity**. The individual members of the domain are defined and named.
Organization
4. Principle of **Differentiation**. The protocol-guided hierarchical structure of a system with categories, and classes within categories.
5. Principle of **Prioritization**. The priority of knowledge by sequencing categories and sequencing classes within categories.
6. Principle of **Diagnostics**. The quantification and use of soil properties, sets of properties, and selected features (diagnostics) that provide objectivity.
7. Principle of **Membership**. Class membership for individuals based on quantified class limits and described central tendencies.
8. Principle of **Certainty**. The recognition that change is inevitable and the driving force continual testing of a system.

C. Attributes of a good soil taxonomic system:

1. Definition of a class or taxon should carry as nearly as possible the same meaning to each user
2. Should be a multicategoric system (hierarchy)
 - a. lower levels - narrowly defined - large number of differentiating characteristics
 - b. higher levels - broadly defined - few soil properties are used to differentiate
3. Classes or taxa should be concepts of real bodies of soil that exist in nature
4. Must be capable of providing for all the soils observed in a landscape
5. Differentiating characteristics should be properties that are observable in the field or quantitatively measured by reliable techniques
6. The system should be capable of modification to incorporate new knowledge with a minimum of disturbance
7. The criteria used should keep undisturbed and cultivated soils in the same taxa.

D. Definition of terms:

1. **Class or taxon** a group of individuals or other units that are similar in selected properties (characteristics) and distinguished from all other classes of the same population by differences in these properties
2. **Population** all the individuals of a natural population collectively
3. **Category** a grouping of classes (taxa) formed at about the same level of classification produced by differentiation within the population being categorized. It includes the entire population.
4. **Pedon** the smallest volume for which one should describe and sample the soil to represent the nature and arrangement of its horizons. It has three dimensions. Its lower limit extends downward to the depth of plant roots or the lower limit of the genetic horizons. Its lateral dimensions are large enough to represent the nature of the horizons and their variability. The minimum lateral cross section ranges from one to ten square meters.
5. **Polypedon or soil individual** it is the smallest natural body that can be defined as a thing complete in itself. It consists of more than one contiguous pedons all of which are within a defined range of soil properties. It is the link between the pedon and the taxonomic unit (the soil series).
6. **Mapping unit** constructed during the course of a soil survey to identify populations of taxonomically identified soil individuals within an area of land. A taxon name is commonly used as part of the mapping unit name.
7. **Soil profile** vertical cut that exposes the horizons for examination or sampling
8. **Soil solum** genetic horizons that have undergone alteration by pedogenic processes
9. **Control section** portion of the pedon used for the family category in Soil Taxonomy



- Key concepts of classification

- Types of characteristics

- **Differentiating**: used to defined classes

- **Accessory**: consistently associated with a class; **co-variant** with differentiating characteristics

- **Accidental**: not associated with the classes

2.1 What is different about soils?

- There is really no soil 'individual' as a self-standing object

- Concept of the **pedon** as a discrete object within the soil **continuum**

- Therefore, the emphasis is on defining **mappable** classes rather than on optimal classification of individuals.

- There is no true inheritance or genetics as it is understood in biology

2.2 Major ways of classifying soils

There are various ways to organise a soil classification. A major distinction is between **natural** and **technical** approaches:

- **Natural** soil classifications group soils by some intrinsic property, behaviour, or genesis of the soils themselves, without reference to use

- **Technical** soil classifications group soils by some properties or functions that relate directly to a proposed use or group of uses.

Natural classifications:

- Group by **ecologic region**, e.g. “prairie soils”, “boreal soils”. Geographically-compact but may have diverse properties and function.
- Group by **presumed genesis**, i.e. the development pathway of the soil profile. These are called **genetic** soil classifications. The soil individual is considered as a natural body with its own history and ecology. This depends on the interpretation of landscape and soil genesis.
- Group by **similar properties**, working **bottom-up** from a set of individuals, to a set of classes, and then grouping the classes into super-classes. This can be done by:
 - Subjective judgement of the classifier
 - Numerical classification, usually multivariate

Technical classifications:

- Hydrologic response [5]
- Suitability classes (FAO Framework for Land Evaluation) [13]
- Land Use Capability (USDA LCC) [20]
- Fertility Capability Classification (FCC) [28] [26]
- Engineering group [23]

SOIL SURVEY

Soil survey

Soil survey is a branch of soil science which involves the identification of the different types of soil in a given landscape and the location of their distribution to scale on a map. In addition, soil survey provides information on the quality of the land in terms of their response to management and manipulation.

From this definition, it is clear that soil classification is a branch of soil survey and the unit of classification is the taxonomic unit or mapping unit or pedon. A taxonomic class depicts the properties of a soil profile as given in the profile description and analyses.

Mapping Unit

A mapping unit is a geographical unit and it is an area of land within which the greater proportion is occupied by the taxonomic class after which it is named. For example, if about 85%

of the soil within a mapping unit is occupied by a soil whose property is that of Ibadan series, the soil will be named as Ibadan series.

Purity of Mapping

The degree of uniformity or heterogeneity in term of kinds of soils within a mapping unit is a measure of its purity. Purity is the proportion of the mapping unit occupied by the profile class after which it is named. It is usually estimated in percentage and different soil survey organizations have different acceptable purity standards. For example, USDA standard is 85% while the Soil survey of England and Wales is 70%.

In accessing the purity of a map, the point used for mapping cannot be used for estimating purity.

Kind of mapping units

Mapping units have been distinguished based on the amount of inclusion or impurity they contain. Five kind of mapping units have thus been distinguished and these are consociation, association, complex, undifferentiated and miscellaneous/unvisited.

Consociation

A consociation is a mapping unit with very little inclusion or impurity. It is assumed to contain the profile class after which it is named but in practice the purity of such class may range from 70% to 85%.

Association

An association is a mapping unit that contain two or more taxonomic class that are nearly equally represented and in which it is very easy to separate one profile class from the other.

Complex

Complex is a mapping unit where more than two taxonomic classes are equally represented and the components are intricately interwoven so that separation, even at large scale is difficult.

Undifferentiated

This is a mapping unit consisting of a number of taxonomic units that are so intricately interwoven that separation into different units are impossible at any reasonable mapping scale.

Miscellaneous/unvisited

This is a loosely used term by some survey organization. It refers mainly to areas that cannot be mapped because of rock outcrops, thick forests or other impediments.

Principles of soil survey

The principles of survey can be discussed under five points

A soil survey must have an objective

The objective or aim of soil survey must be defined before the commencement of the survey. It may be wide ranging such as to provide a relatively stable data base that will last for many years and be useable for a variety of purposes, or it may be narrow and specific such as to delineate the land suitable for irrigation in a particular area.

A soil survey is not the only basis for decision on land use and management, it is only an aid

Decisions on land management are usually influenced by economic considerations, social and institutional factors, often by existing legal land rights and sometimes by political constraints. Even on the more limited sphere of the physical environment, the soil is only one factor. Slope angle, frequently has a dominant effect on the choice between arable and non-arable use and climate is the main determinant in the choice of crop. The findings of soil survey and land evaluation are usually guides to land use.

Land resources do not consist of soils alone

The potential of land to support crops depends on climate as much as on soils, and whenever soil-water relationship can be advantageously modified. Also, applied ecological surveys are more important than soils as a basis for livestock production and extractive forestry. Landform and characteristics of the deeper regolith are frequently more important to the engineers than the upper 1-2 meters studied by the soil surveyor. Therefore, when commissioning or considering a soil survey, one should bear in mind the other kinds of natural resources information that may be necessary and whether these can be incorporated into the soil survey or will require other specialist's investigations.

A soil map must show soils

The map produced by a survey is a soil map if mapping units are based specifically on soil profile. Many surveys rightly include features of land forms in the description of the mapping units. A map based on classes such as "soil-land form association" is a soil map if it is directed towards showing the distribution of soils; the land forms being used as a means to an end. If on the other hand, it is primarily a map of land form units with soils being added to the legend, then it is a geomorphological map.

Soil map and report are complementary

The products of a soil survey include a soil map and survey report; neither is more important than the other because they are mutually indispensable. The amount of information that can be printed on the face of a map is limited and must be supported by data given in the soil survey memoir or report. The report is however, more than just an amplified legend. It contains in addition, background information on other environmental factors, information on land potential and probable response to various alternative forms of management. And sometimes also land use management recommendation. One soil survey report may serve several map sheets.

Kinds of Soil Survey

Soil survey have been distinguished on the basis of three criteria:- purpose of survey, regularity of observation and scale of mapping (intensity).

Based the purpose of the survey (What do you want to do with the soil survey report?), there are two (2) types of surveys:- special purpose and general purpose surveys.

A general purpose soil survey is one that is done mainly to add to the already existing inventory of soil information. This commonly found in the national survey of each country, e.g. the USDA, FMAWR. The information may not be needed at the time of survey but such a survey is done for record purpose. Such survey usually employs many differentiating properties so that it may be found useful for several purposes that may arise in the future.

A special purpose soil survey is done for specific purpose in mind, e.g. survey for irrigation or survey for citrus plantation. In such survey, the properties (differentiating characteristics) of the land which is important for the purpose are emphasized. Special purpose surveys are usually done at large scale or semi-detailed. This has to be so because the area covered are smaller and also the intensity of observation also need to be higher.

Based on regularity of observation, three kinds of surveys have been distinguished: - free survey, rigid grid and flexible grid.

In **free survey**, there is no rigid pattern of observation. The surveyor uses the field features such as change in vegetation, topography, slope and even change in sound to movement to observe soil and to locate soil boundaries. Usually there is a lot of remote sensing methodology involved in free survey. And in most modern soil survey involving 5000 ha and above, this is usually the type of survey methodology adopted. The advantage is that it is cheaper because there is less

number of observation points. However, because it depends largely on the experience of the surveyor, it can be very inaccurate with an amateur, since boundary placement is more difficult.

In **rigid grid survey**, examinations of the soil are done at regular and pre-determined interval. It is normally used when detailed information is required, e. g. mapping the soil of a research station or mapping for irrigation. Usually, the points of observation are at the intersection of the two regularly placed vertical and horizontal lines.

It has an advantage in thickly forested area where visibility of the terrain is poor. It is usually not used or recommended in large areas as it is expensive and the accuracy may not deserve the extra cost. It is the best method of survey for amateurs.

Flexible grid survey method is a compromise between the free and rigid grid methods of survey. In this system of survey, the number of observation is fixed but the location of the observation points are not pre-determined and can be fixed at will.

Based on the scale of mapping, there are seven kinds of surveys:- compilation, integrated survey, exploratory survey, reconnaissance survey, semi-detailed survey, detailed survey and intensive survey.

Compilation: These are soil maps produced by abstraction from other soil surveys. And where they exist they are filled by inferences. The scale is usually at 1: 100,000 or smaller. Many national soil maps of many countries are produced in this way.

Integrated survey: This is also known as land system survey. It is based on mapping the total physical environment and in fact land forms are mapping unit. Soils are an important but usually not a defining property of the mapping unit. The scale is 1: 250,000 or smaller.

Exploratory survey: Exploratory surveys are not survey proper. They are usually rapid road traverse made to provide modicum of information about the area that are otherwise unknown. Scale of exploratory survey varies from 1: 2,000,000 to 1,500,000.

Reconnaissance survey: These are mostly based on remote sensing especially Area Photo Imagery (API). They are the smallest scale of survey where the whole area is still covered. The scale is usually 1:250,000 although smaller scales have been used.

Semi-Detailed survey: In a semi-detailed survey, we have a combination of remote sensing and field work. Mapping units are usually soil association. Scale of mapping varies from 150,000 to 100, 000.

Detailed survey: Detailed surveys are executed through field examination with pre-determined numbers of observation points and or spacing. These kinds of surveys are usually employed for small area and for special purposes. Scale of observation varies between 1: 10,000 and 1: 25,000. Mapping unit are usually soil series.

Intensive survey: Intensive survey rigid grid approach, i.e. number of observation and spacing of observation are pre-determined. Mapping units are soil series and phase of soil series. Scale of mapping varies from 1: 1,000 to 1: 10,000 or even larger. They are usually experimental station surveys.

PART B: SOIL PHYSICS

DEFINITION OF SOIL PHYSICS

Soil Physics is a branch of soil science that deals with physical properties of soil as well as measurement, prediction and control of physical processes taking place in and through the soil. Soil physical properties include soil texture, soil structure, soil colour, consistency, density thermal regime, soil water, porosity, infiltration, hydraulic conductivity etc.

Soil Physics could fundamentally be regarded as both basic and applied sciences. This is because, Soil Physics involves application of the principles of Physics to the characterization of soil properties and understanding of soil processes involving transport of matter or energy.

SOIL PRODUCTIVITY

Soil productivity is an economic concept and signifies the capability of the soil to produce specified plant or sequence of plants under well defined specified systems of management and environmental conditions. This suggests that productivity is not soil fertility alone but a function of several factors (e.g. climatic condition and soil factors). Soil productivity is measured in terms of output or harvest.

On the other hand, soil fertility refers to the inherent capacity of the soil to provide adequate amount and proper balance of nutrient for the growth of specified plant when other growth factors (e.g. light, water, temperature and favourable soil physical environment) are favourable.

In addition to chemical fertility i.e. presence of adequate nutrient in the soil and absence of toxic agents, the soil should also, be physically fertile. That is, the soil must be loose, soft and friable, possesses no mechanical impedance to root development, has pore volume and size distribution that allow entering, movement and retention of water and air to meet plant needs and has optimal thermal regime.

SOIL AS A DISPERSED SYSTEM

Soil is made up of 4 basic components: mineral matter; organic matter, soil water and soil air.

On the basis of these, there are three phases in the soil. These are solid phase, liquid phase and gaseous phase.

The Solid Phase: The solid phase is broadly composed of inorganic (mineral) and organic constituents. The mineral constituents form the bulk of the soil solid phase and consist of particles of various sizes, shapes and chemical composition. The mineral constituent specifically composed of primary and secondary minerals. The primary minerals are the quartz and feldspars (most abundant) with relatively small amount of pyroxenes, amphiboles, olivine, micas etc.

The secondary minerals originates from the break down of primary minerals and examples are silica, alumina, iron oxides etc. They constitute the most active site in the soil.

The organic fraction constitutes a small fraction of the soil solid phase and it includes residues at various stages of decomposition as well as life organisms.

The Liquid Phase: About 40 to 50% of the bulk volume of soil body is generally occupied by soil pores or voids which may be partially or completely filled with water. The liquid phase is an aqueous solution of salts because the soil water keeps salts in solution which act as plant nutrients. When all the soil pores are completely filled with water, the soil is said to be saturated.

The Gaseous Phase: This is the soil air/atmosphere. It is made up of mixture of gases. It composed mainly of nitrogenous gas, oxygen, carbon dioxide and water vapour. The volume of the gaseous phase is dependent on that of the liquid phase. The pore space not filled by water is automatically occupied by air.

The dispersed nature of the soil and its constituent inter-phasal activities give rise to such phenomenon as:

- (i) **adsorption of water and chemicals;**
- (ii) **ion exchange;**
- (iii) **adhesion and cohesion;**
- (iv) **dispersion and flocculation;**
- (iv) **swelling and shrinking and**
- (v) **capillarity.**

The three phases of the soil play definite roles. The solid phase provides mechanical support for and nutrients to the plants. The liquid phase supplies water and along with it dissolved nutrients to plant root. The gaseous phase satisfied the aeration need of the plant. Thus, the 3 phases complimentarily shared the soil's function to sustain plant growth.

SOIL TEXTURE

Soil texture is the relative proportion of various soil separates in a soil. It is usually expressed on percentage basis.

Soil separates are group of soil particles of given size range i.e. different size of particles which together make up a given soil.

The main textural classes are sand, silt and clay. These textural classes may be modified by addition of suitable adjective based on relative amount of each separate that make up the soil e.g.

Loam: Soil material with clay, silt and sand in close proportion (e.g. 7-27% clay; 28-50% silt and <50% sand).

Loamy sand: Materials with about 80-90% sand.

Sandy loam: <7% clay; <50% silt; about 52% sand.

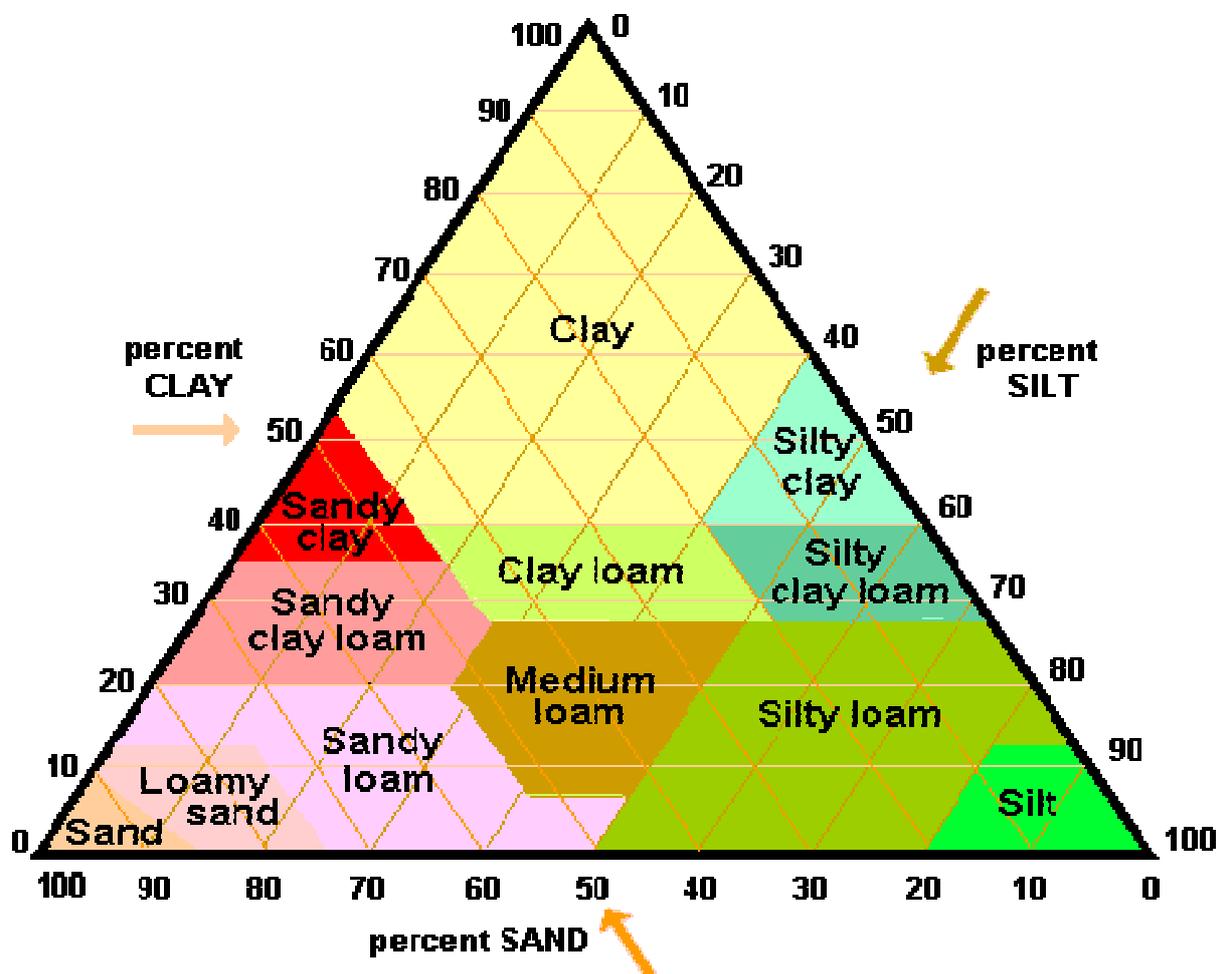
Other modifications include silty loam, sandy clay loam, clay loam, gravelly loamy sand etc.

Determination of Soil Texture

Soil texture may be determined on the field by textural feel and in the laboratory by soil mechanical analysis or soil particle size distribution. The mechanical analysis in the laboratory may be carried out either by Pipette or hydrometer method.

After the proportion of each of the soil separates are determination, the textural class of the soil is identified using a USDA Soil Textural Triangle. The sides of the soil texture triangle are scaled for the percentages of sand, silt, and clay.

Textural Triangle:



Systems of soil particle size classification

There are two widely used systems of soil classification. These are: United State Department of Agriculture (USDA) and International Soil Science Society (ISSS)

USDA Classification system

<u>Fraction</u>	<u>Diameter (mm)</u>
Very coarse sand	2.00 – 1.00
Coarse sand	1.00 – 0.50
Medium sand	0.50 – 0.25
Fine sand	0.25 – 0.10
Very fine sand	0.10 – 0.05
Silt	0.05 – 0.002
Clay	<0.002

ISSS Classification system

<u>Fraction</u>	<u>Diameter (mm)</u>
Coarse sand	2.00 – 0.2
Fine sand	0.2 – 0.02
Silt	0.02 – 0.002
Clay	<0.002

Generally

Materials: >20 mm diameter – stone

20-2 mm diameter – gravel

<2mm diameter – Fine earth (soil)

Importance of soil texture

- It affects water and nutrient holding capacity of the soil
- It influences the type of crop to be grown
- It indicates type of management needed for crop growth and for engineering purposes.
- **Mechanical composition of soil**
- The mechanical composition of soil is a basic requirement in the soil physical investigation useful for land capability classification and in the study of soil morphology, genesis, classification and mapping.
- Soil mechanical analysis is the procedure for determining the particle size distribution of a soil sample.

Steps in soil mechanical analysis

- ❖ Sample collection
- ❖ Air dry the sample at room temperature
- ❖ Dispersion of the sample in an aqueous solution using Calgon solution (Sodium hexametaphosphate).
- ❖ If the sample contains high amount of organic matter remove the organic matter using H_2O_2 .
- ❖ Carry out mechanical agitation by shaking or using ultrasonic vibration

❖ Determination/quantification of size fraction

(1) Sieving (for coarse fraction) – use net of sieve corresponding to the desired particle size

(2) Sedimentation (fine fraction)

The principles of sedimentation are that the velocity of fall of particle in a viscous medium is influenced by

- (i) the viscosity of the medium
- (ii) density difference between the medium and the falling particle
- (iii) the size and shape of the material.

The law which govern sedimentation of particles is called Stoke's Law which states that resistance offered by liquid to the fall of a rigid spherical particles vary with the circumference of the sphere (and not its surface) **OR** the terminal velocity of a spherical particle settling under the influence of gravity in a fluid of a given density and viscosity is proportional to the square of the particle radius.

The Stokes' law consists of the factors contributing to the cause of settling and resistance to settling.

Particle volume = $\frac{4}{3}\pi r^3$; density difference = $d_1 - d_2$; g = gravity; Particle circumference = $2\pi r$

Viscosity = η ; V = velocity of sedimentation

Therefore, if

Force of settling = resistant to settling

Force = mass x acceleration i.e $F = mg$ (where g is acceleration due to gravity)

Note that density = mass/volume; therefore mass = ρv

$$\frac{4}{3}\pi r^3 (d_1 - d_2)g = 2\pi r \cdot \eta \cdot 3v$$

$$V = \frac{4}{3}\pi r^3 (d_1 - d_2)g = \frac{2r^2(d_1 - d_2)g}{9\eta}$$

$$2\pi r \cdot \eta \cdot 3$$

$$9\eta$$

V = settling velocity (cm s^{-1}); d_1 = particle density; d_2 = density of fluid (g cm^{-3});

r = particle radius (cm); η = viscosity of the medium

The time (t) required for particle to settle through a given depth (h) in liquid may be obtained from the Stokes' Law

$$V = \frac{2r^2(d_1 - d_2)g}{9\eta}$$

$$9\eta$$

Re-write r^2 in terms of particle diameter (D) i.e. $r^2 = \frac{1}{4} D^2$

Therefore, $V = \frac{2D^2(d_1 - d_2)g}{36\eta}$

$$36\eta$$

$$V = \frac{(d_1 - d_2)g D^2}{18\eta}$$

$$18\eta$$

Velocity = distance (h) / time (t); hence, $t = h/V$

$$t = \frac{18\eta \cdot h}{(d_1 - d_2)g D^2}$$

$$(d_1 - d_2)g D^2$$

Example:

Calculate the velocity of sedimentation of silt particle 0.006 mm and clay particle 0.0002 mm diameter at 30°C. The viscosity of the water at 30°C was 0.00798

$\text{cm}^{-1} \text{s}^{-1}$. Assuming the gravity to be 980 cm s^{-2} .

SURFACE RELATIONSHIP

The extent of the surface of dispersed soil system is described in terms of the soil specific surface.

The soil specific surface is defined as the sum of the surfaces of constituent dispersed soil particles referred to unit mass or unit volume.

Specific surface of soil (A_m) or (A_v) = Total surface area of soil (A_s)

Mass or volume of soil (M_s or V_s)

That is, A_m or $A_v = \frac{A_s}{M_s \text{ or } V_s}$

M_s or V_s

Therefore, on mass basis $A_m = \frac{A_s}{M_s} \frac{\text{cm}^2}{\text{g}}$ or $\frac{\text{m}^2}{\text{g}}$ or $\frac{\text{m}^2}{\text{kg}}$

M_s g g kg

On volume basis, $A_v = \frac{A_s}{V_s} \frac{\text{cm}^2}{\text{cm}^3}$ or $\frac{\text{m}^2}{\text{m}^3}$

V_s cm^3 cm^3

Many of the soil physical and chemical reactions are related to inter facial surface phenomena and are thus influenced by soil specific surface area.

Soil properties such as plasticity, swelling, soil strength, water retention, CEC and nutrient availability are strongly affected by soil surface area.

Soil specific surface depends on:

- i. Particle size
- ii. Particle shape
- iii. Mineralogy of the materials

VOLUME-MASS RELATIONSHIP

The volume-mass relationship among the three phases of soil could be diagrammatically represented as:



Based on the diagram above, a quantitative representation of the 3 phases can be expressed in terms:

$$\text{Particle density } (\rho_s) = M_s/V_s \text{ (g cm}^{-3}\text{)}$$

V_s = volume of solid

$$\text{Bulk density (dry): } (\rho_b) = M_s/V_t \text{ (g cm}^{-3}\text{)}$$

$$\text{Total porosity } (P_t) = V_f/V_t \text{ or } (V_a + V_w)/V_t$$

$$\text{Void ratio } (e) = (V_a + V_w)/V_s$$

Soil wetness: This can be expressed relative to total mass or total volume

In term of mass: mass wetness (w) = M_w/M_s . This is called gravimetric water content

In terms of volume (Θ) = V_w/V_t . This is called volumetric water content.

$$\text{Degree of saturation } (s) = V_w/V_f$$

Air filled porosity (P_a) = V_a/V_t . This is the fractional volume of air in the field.

SOIL STRUCTURE

Soil structure is the arrangement of soil particles to form peds. Or, the arrangement of primary particles into secondary particles (aggregate).

Soil structure is strongly affected by changes in climate, biological activities and soil management practices.

Measurement of soil structure

There are direct and indirect methods of measuring soil structure.

The direct method involves measuring the size and shape of aggregate and pores of the soil. That is, three dimensional study of soil. This is done by thin section analysis with aid of powerful microscope e.g. scanning electron microscope, transmission microscope, petrographic microscope etc.

The indirect methods involve measuring soil properties that depend on soil structure. These properties include: aggregate size distribution, aggregate stability, bulk density, porosity, pore size distribution, permeability, infiltration etc.

Importance of soil structure

- It affects water and nutrient holding capacity of the soil
- It affects germination and root growth and development
- It affects water retention and transmission of fluid in soil
- It affects soil aeration
- It influences soil thermal properties

Measurement of soil structural stability

There are various methods: the two common methods that will be considered are sieving method and water drop impact technique

1. **Sieving method:** there are wet and dry sieving techniques. What is measured here is the Mean weight diameter (MWD). For wet soil it is called MWDW and for dry soil it is called MWDD

The formula is given as

$$MWD = \sum_{i=1}^n X_i W_i$$

Where X = mean of diameter of particles size range separated by sieve

W_i = weight (proportional) of the aggregates in each size range

Example

Aggregate diameter (cm)	%	X	W _i	X W _i
10 – 5	5	7.5	0.05	0.375
5 – 2	5	3.5	0.05	0.175
2 – 1.0	15	1.5	0.25	0.225
1.0 – 0.5	25	0.75	0.25	0.1875
0.5 – 0.0	50	0.25	0.5	0.125
			MWD	1.087 mm

2. **Water drop technique:** this is done by studying the impact of rain drop on soil aggregate stability. This is done in the laboratory by using a rainfall simulator.

SOIL WATER

Soil water is the amount of water present in the soil available to crop. Water can enter into the soil by precipitation or by irrigation.

Forms in which water exist in the soil

Water exists in 3 forms in the soil

(i.) Hygroscopic water (ii.) Capillary water (iii.) Gravitational water.

Importance of water to crop

- i. It is important in the absorption of mineral salts from the soil by plant.
- ii. It helps in the transportation of plant nutrients from the root to other parts of the plant.
- iii. Water is an essential raw material needed during photosynthesis.

Some terminologies in soil water include

Field capacity, water table, surface tension, water log, wilting point

Ways by which soil loss water

- i.) evaporation from the soil surface
- ii.) Transpiration from plant leaves, stem and fruit surface.
- iii. drainage
- iv.) erosion

Sources of water in the soil

i.) Precipitation

ii.) Irrigation

iii.) High humidity

Methods of determining soil water content

1. Gravimetric method
2. Volumetric method
3. Neutron scattering technique
4. Gamma ray attenuation technique
5. Electric resistant method
6. Time domain reflectometry
7. Tensiometry method

Method of expressing soil water content

1. On mass basis
2. On volume basis
3. On depth basis

Forces acting on soil water

1. Matric force
2. Osmotic force
3. Body force

Movement of water in soil

Flow of water in soil may take the following forms:

1. Saturated flow
2. Unsaturated flow
3. Vapour movement

Soil Air and Aeration

- **Soil air** - is the air that fills the soil pore spaces not occupied by water. The gaseous phase of the soil not occupied by solid or liquid.
- **Soil aeration** - is the process of exchange of oxygen and carbon dioxide of soil air with the atmosphere.

Soil Air Composition

Soil Air Composition

Gas	Atmosphere (%)	Soil air (%)
Oxygen	21	10 to 20
Carbon dioxide	0.03	0.10 to 5
Nitrogen	78	78.8 to 80
Argon	0.94	
Hydrogen	0.01	

Factors Influencing Soil air Composition

- Organic matter content
- Microbial activity

- Plant root respiration
- Texture
- Structure
- Water content
- Crops
- Drainage
- Tillage
- Season

Soil Air Capacity

Definition:

- Is the fractional volume of air in the soil at field capacity.
- That is the quantity of air in the soil after soil has been saturated and allowed to drain for about twenty-four (24) hours.

Poor Soil Aeration

Poor soil aeration has adverse effects which could result in certain changes that are –

1. Morphological
2. Physiologic
3. Anaerobic conditions in soil induce series of reduction reactions

Soil Thermal Properties

Soil thermal regime determines:

1. The rates and directions of some physical processes.

2. The rates of energy and mass exchange with the atmosphere.
3. Governs the type and rates of chemical reactions taking place in the soil e.g. weathering.
4. Influences biological processes such as:
 - i) Microbial activity
 - ii) Soil germination
 - iii) Plant growth

Major Aspects Of Soil Thermal Regime

The major aspects which characterize the soil thermal regime are:

1. Soil heat intensity: This describes mainly the soil temperature ($^{\circ}\text{C}$, degree Celsius; K, Kelvin).
2. Soil heat capacity (gravimetric): The amount of heat required to raise the temperature of a given mass by one degree Celsius (1°C)

Mode of Heat Energy Transfer in Soil

Mode of heat energy transfer in soil include:

- i) Conduction – This is the primary mode of heat transfer in soil.
- ii) Convection
- iii) Radiation

Factors Influencing soil Temperature Variation

1. Factors that influence the amount of heat available at the soil surface are:
 - i) Soil colour
 - ii) Soil mulch

2. Factors that influence dissipation of available heat

i) Water content of the soil

Types of Soil Temperature Variation

1. Diurnal
2. Seasonal
3. Variation due to soil depth

Solar Radiation

Solar Radiation is the major source of soil heat. Only a portion of the emitted solar radiation reaches the earth's surface. Part of the solar radiation may be:

1. Reflected by the clouds
2. Scattered into the atmosphere by atmospheric gases.
3. Absorbed by the ozone and water vapour.

Management of Soil Heat

These methods includes:

1. Covers
2. Mechanical manipulation of soil surface
3. Others (indirect effects)
 - i) Irrigation – reduces temperature
 - ii) Drainage – increases temperature
 - iii) Weed control

- iv) Plants/trees

SOIL EROSION

Definition

Soil erosion can be simply defined as the wearing away of soil materials from place to place by the agents of erosion such as water, wind and ice.

In general soil erosion is broadly divided into

1. Geological Erosion

Soil erosion that occurs naturally, without the influence of human activities.

2. Accelerated Erosion

Soil erosion resulting from human interference with the natural environment.

Mechanics of Soil Erosion

- a. **Detachment** of soil aggregates into particles
- b. **Transportation** of the detached particle by floating, rolling, and dragging.
- c. **Deposition** of the transported materials where the energy of force dissipates.

Soil erosion by water and wind erosion involves the three processes listed above. However, the method of soil movement in wind erosion defers.

Factors causing soil erosion

- 1. Climatic factor

2. Soil factors
3. Topography
4. Vegetation cover
5. Human activities e.g., *Tillage, Overgrazing, Fires, Lowering of the water table (water use in excess of replenishment rate)* - these accentuates wind erosion

Types of Soil Erosion by water

- **Sheet Erosion**

The removal of a fairly uniform layer of soil from the land surface by runoff water.

- **Rill Erosion**

As sheet flow is concentrated into tiny channels (*called rills*), rill erosion occurs.

- **Gully Erosion**

When the volume of runoff is concentrated, the rushing water cuts deeper into the soil, deepening and fusing the rills into larger channels called gullies.

Measurement of Soil Erosion by Water

The **Universal Soil Loss Equation** (USLE), was designed by Wischmeier and Mannering, (1969), to predict annual soil loss by water in the USA but has been adapted and modified in some cases for prediction around the world. The USLE equation is as follows:

$$A = R K L S C P$$

A - predicted soil loss ($\text{kg m}^2 \text{s}^{-1}$)

R - rainfall erosivity

K - soil erodibility

L - slope length

S - slope gradient or steepness

C - cover and management

P - erosion control practices

Wind Erosion

Types of soil movement

1. Saltation
2. Soil Creep
3. Suspension

Factors affecting Wind Erosion

1. Wind velocity
2. Soil characteristics / properties
3. Vegetation / mulch